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Bottom boundary layer flow and salt injection from the continental shelf to slope

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[1] Austral winter oceanographic measurements from the northwest Australian continental shelf reveal salty water forming evaporatively inshore, moving across the wide shelf near the bottom and into the adjacent open ocean when the shelf edge alongshore flow is equatorward. The salt tongue is absent during more normal conditions, when the poleward Leeuwin Current is present. We hypothesize that the flow reversal enables shelf-wide bottom boundary layer (Ekman) transport and thus creates the shelf-edge convergence that accounts for the observed salt tongue. This flow is absent under sustained normal conditions because of buoyancy arrest in the bottom boundary layer.

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1. Introduction

[2] Cross-shelf exchanges have not previously been intensively studied over the wide (typically 150 km) shelf off the northwestern coast of Australia. The dominant sustained coastal current here is the southwestward-flowing (poleward) Leeuwin Current, which is strongest (0.3–0.4 m/sec) near the shelf edge and continues to flow southward along the western coast of Australia [*Church and Craig*, 1998], ultimately joining the South Australian current [*Ridgway and Condie*, 2004] to form a continuous feature more than 5000 km long. The Leeuwin Current is centered near the 100–150 m isobaths and is generally associated with a core of anomalously fresh water [*Holloway*, 1995]. The northwestern Australia region is interesting not only because it is the headwaters of a long current system, but also because of its highly energetic tides, associated nonlinear internal waves [*Holloway*, 1995], weak wind forcing in the absence of tropical cyclones [*Church and Craig*, 1998], and strong evaporative fluxes [*Holloway*, 1995].

[3] During June–July 2003, we conducted a month-long research cruise (Figure 1) to study the role of evaporative densification over the inner shelf and the anticipated hydrodynamical instabilities on the Leeuwin Current core (with consequent implications for shelf-ocean exchange). The densification problem was particularly intriguing since this process is normally associated with wintertime cooling at high latitudes, where observing conditions are much less convenient. Our measurement program concentrated on direct current observations and progressively better-resolved temperature and salinity mapping (http://science.whoi.edu/users/seasoar/). Sampling began on a relatively broad spatial scale and was increasingly concentrated on the shelf near Port Hedland, where a single current meter mooring was placed near the outer shelf to monitor the Leeuwin Current.

2. The Observations

[4] During our sampling period, shelf-edge alongshore currents begin in the normal poleward Leeuwin Current sense, reverse for about 9 days, and finally return to poleward flow (Figure 2). The shelf-edge current reversal coincides with a sudden drop in coastal sea level, suggesting this is not just an isolated feature, but a shelf-wide phenomenon, apparently associated with wind driving. Before the reversal, the hydrographic structure (measured by 9 cross-shelf sections distributed over an alongshore scale of about 600 km) is consistent with previous observations [*Holloway*, 1995] in that the shelf edge flow is poleward and has a relatively fresh core (Figure 3). Cooler, saltier water, consistent with evaporative effects (we measured mean net evaporation to be 0.17 m/month), is found closer to shore, but it does not extend offshore beyond about the 100 m isobath. During this period, shelf waters are saltier toward the southwest, with an alongshore gradient of about 6×10^{-7} 1/m. On around June 29, however, the shelf-edge alongshore flow reverses to northeastward (equatorward) throughout the water column and hydrographic conditions change radically (Figure 4). Specifically, the saltier water is no longer confined to the shelf proper: it runs offshore along the bottom, separates near the shelf edge, and extends 15–45 km offshore into the deep ocean, roughly along the 23.5 isopycnal. The tongue waters are about 0.2 saltier than ambient waters, a salinity change consistent with a relatively small density (σ_t) perturbation of about 0.2.

[5] During the 9-day period of equatorward shelf edge flow, we collected 21 synoptic, highly resolved cross-shelf hydrographic sections, and all show salinity tongues comparable to that in Figure 4. Our shipboard, underway acoustic Doppler current measurement system could not sample the bottom 15 m of the water column, and oceanic "noise", such as internal tides, is sufficient to make the relatively weak velocities associated with the salt tongue unobservable. The mean alongshore flow during the flow reversal period (Figure 5) is clearly poleward (negative) and relatively depth-independent inshore of the 125 m isobath, but equatorward just offshore of the shelf. During the reversal, the fresh water core associated with the Leeuwin

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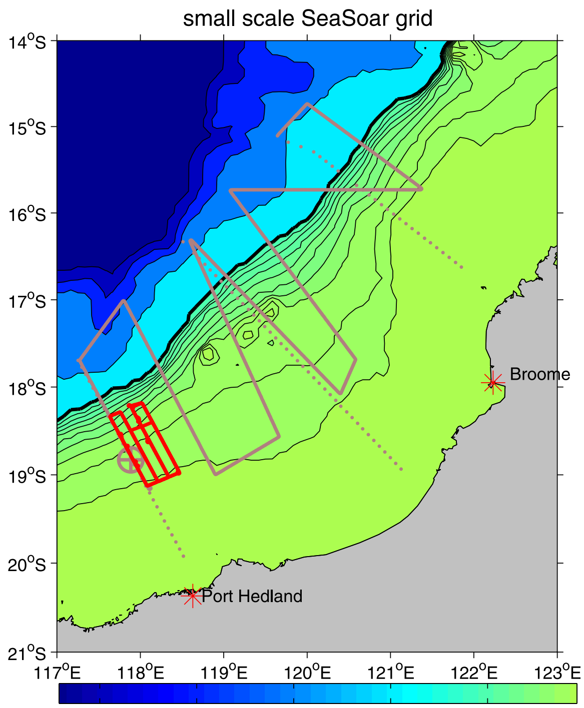

Figure 1. Locator map of the study region. Depth contours are at 100 m increments until the 1000 m isobath, and then occur at 1000 m increments in deeper water. The 1000 m isobath is a heavy black line separating green from blue colors. The dotted lines show the locations on CTD stations, the solid red lines are the location of repeat SeaSoar underway sampling, and the brown line shows the location of large-scale (one-pass) SeaSoar sampling. The circled cross is the location of the current meter mooring.

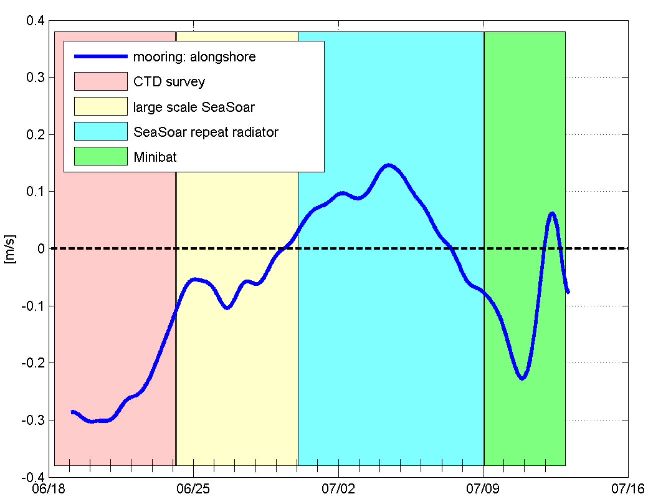

Figure 2. Depth averaged, low-pass filtered (tides removed) time series of alongshore velocity at the moored current meter location. LP filtered time series at the ADCP mooring, with the repeat radiator time shaded. Background colors denote the type of sampling underway at a given time.

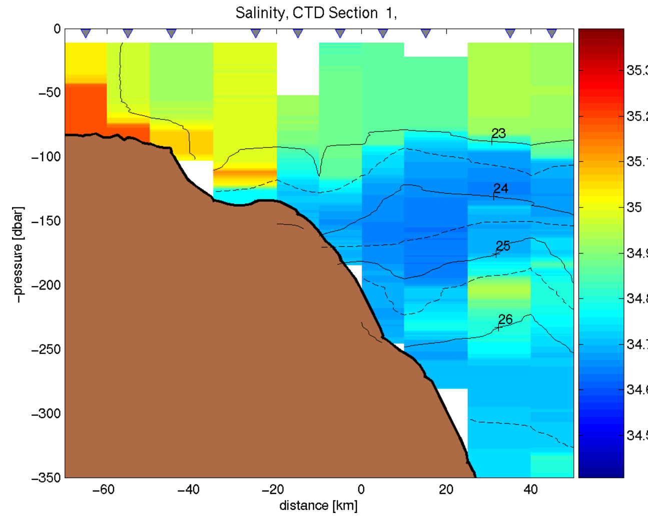

Figure 3. Cross-shelf/vertical section of salinity (color) and density (solid contours) from CTD section A (westernmost dotted line of Figure 1), June 18–20, 2003.

Current is found offshore of the equatorward jet. Our measurements, along with older observations [e.g., *Holloway*, 1995] lead us to conjecture that the high salinity tongues are consistently found when sustained alongshore flow over the outer shelf is reversed relative to its normal southwestward condition, but not present otherwise.

3. The Mechanism

[6] Scaling of the cross-shelf momentum equation in the bottom boundary layer shows that small density variations associated with the saline tongues are probably less important than the bottom stress in this context, so we do not believe that density-driven cascading accounts for the observed cross-shelf movement. Rather, we think that transport in the bottom Ekman layer [e.g., *Gill*, 1982] is the key process. Simply stated, in idealized homogeneous conditions, the bottom stress associated with an alongshore flow must give rise to an orthogonal, cross-shelf transport of

$$U_E = -\tau^y/(\rho f) \tag{1}$$

(where τ^y is the alongshore bottom stress, ρ is the water density and f is the Coriolis parameter) in the turbulent bottom boundary layer, which is expected to be of order 10–20 m thick. The transport is to the right of the overlying flow in the Southern Hemisphere. The strength of this transport is proportional to the bottom stress, which in turn relates nonlinearly to the strength of the alongshore flow. Convergence (divergence) in the bottom boundary layer transport must be compensated by upward (downward) motion near the bottom. Thus, if the alongshore flow reverses as a function of cross-shelf distance, it implies that there will be water either added to or withdrawn from the bottom boundary layer.

[7] Modern bottom boundary layer theory [e.g., *Trowbridge and Lentz*, 1991], has shown that Ekman transport over a slope is constrained by the vertical density stratification. After an advective adjustment time,

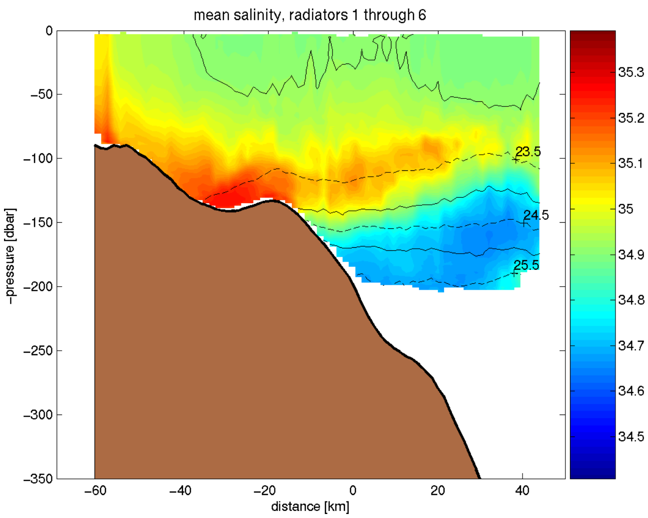

Figure 4. Cross-shelf/vertical section of salinity (color) and density (solid contours) from the average of all 21 cross-shelf sections in the repeat sampling area (red grid on Figure 1), June 30–July 7, 2003.

cross-isobath density gradients develop in the bottom boundary layer, neutralizing the bottom stress because of geostrophically balanced shear in the lower tens of meters. Under idealized circumstances, this neutralization, or "buoyancy shut-down", happens over a time scale [*Garrett et al.*, 1993] of

$$T_B = (f/\alpha N)^3 V/(rN), \tag{2}$$

during offshore flow (where α is the bottom slope, N^2 the squared buoyancy frequency $-g\rho^{-1}\,\partial\rho/\partial z$, V is the ambient alongshore velocity, ($\rho\ r$) is the proportionality constant between the alongshore bottom velocity and the bottom stress (and r has dimension of velocity), and g is the acceleration due to gravity). The shutdown is generally quicker when bottom layer flow is onshore than when the flow is offshore [*MacCready and Rhines*, 1993; *Ramsden*, 1995]. The shut-down time is critical when considering cross-shelf transport in the bottom boundary layer because long time scales allow more net cross-shelf transport before bottom stress ceases, while short shut-down times will rapidly arrest that transport.

[8] Now consider the case over the northwestern Australian shelf, during alongshore flow reversal (Figure 5), when the shelf-edge flow is toward the northeast (equatorward) and currents over the shelf proper remain southeastward (poleward). In the absence of shut-down effects, bottom boundary layer flow beneath the shelf-edge jet is onshore, but over the shelf proper, bottom boundary layer flow is offshore. Thus, a convergence occurs centered on about the 150 m isobath, where alongshore flow changes direction. For the conditions of June 29–July 7, 2003 over the shelf inshore of the jet, T_B is about 30–50 days. Under the shelf-edge jet, stratification is stronger than on the shelf and the bottom slopes more steeply, so the shut-down time is much shorter, around 1 day. Thus, over the nine days after the flow is perturbed by the shelf-edge Leeuwin Current reversal, the flow coming off the shelf is largely unaffected by buoyancy shut-down, and salty inshore water is moved O(50 km) across the shelf, but the onshore flow under the jet is likely arrested quickly. The convergence near the 150 m isobath remains, and the ejected water is predominantly salty shelf water, rather than a mixture of the salty water with fresher slope waters.

[9] Ekman convergence implies an injection of bottom boundary water upward onto isopycnal surfaces in the interior, and in our case these injected waters are "dyed" with a higher salinity from inshore. The process is taken to be approximately uniform in the alongshore direction, so that the ejected salty water, if it is constrained to stay on or near constant density surfaces, must then mix and/or flow offshore. The offshore transport is expected to be of order U_E if mass is conserved. It thus appears that the expulsion of water from the bottom boundary can give rise to substantial offshore interior transports across the shelf-ocean boundary. We were able to observe this process conveniently here because evaporative processes much closer to shore "tagged" the bottom boundary layer water with a relatively easy-to-measure salinity marker.

[10] During typical Leeuwin Current conditions, offshore extension of salty water is evidently not observed. One consistent explanation for this is that the poleward alongshore flow over the shelf proper persists for longer than the buoyancy shut-down time (30–50 days), and the cross-shelf transport in the bottom boundary layer is arrested. Thus, salty inshore water is contained over the shelf as in our initial observations. If this is the case, and evaporation continues over the shelf, the resulting salty water inshore of the 100 m isobath must then be advected poleward so that water becomes saltier toward the southwest. The resulting simple salinity balance, that alongshore advection balances the surface evaporative salinity source, would hold if the alongshore flow is in the range of 0.04–0.15 m/sec,

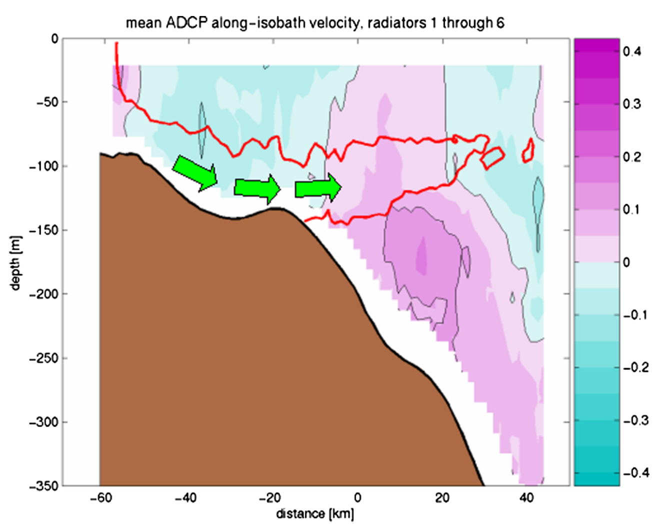

Figure 5. Averaged (over the same sections as Figure 4) alongshore velocity (m/sec), measured from the ship's underway acoustic Doppler device (color and black contours). Flow is positive toward the northeast. The heavy red contour represents the 35.0 isohaline from Figure 4. The heavy green arrows represent, schematically, the proposed cross-shelf flow in the bottom boundary layer, and how it separates.

depending on the water depth. These are physically sensible velocity numbers and are comparable with our direct measurements. Thus, we conjecture that extended periods of poleward flow bring cross-shelf Ekman transport to a halt.

[11] The key ingredients to our hypothesized salt tongue mechanism are 1) the flow reversal that represents a disruption of the typical Leeuwin Current state so that buoyancy arrest is not initially effective, 2) the long buoyancy shutdown time (due to weak stratification) over the shelf which allows bottom Ekman transport to continue during the entire 9 day reversal period, and 3) the convergence created by the flow reversal at the shelf-edge which injects the salty boundary layer waters into the water column. We expect to find these ingredients on the many continental shelves where a) buoyancy losses (which can occur through either evaporation or through wintertime cooling) homogenize the shelf water column, thus inhibiting buoyancy arrest over the shelf, and b) unsteady conditions, which can readily occur as a result of wind forcing [e.g., *Brink*, 1998], reinitialize offshore bottom Ekman transport.

4. Implications

[12] We have observed a relatively unexplored means of transporting coastal waters offshore. The process is somewhat similar to one proposed for frontal regions [*Gawarkiewicz and Chapman*, 1992], and observed briefly using intentional tracer releases, south of New England [*Houghton and Visbeck*, 1998]. The Australian example is similar to wintertime conditions south of New England, in that the shelf waters are weakly stratified and thus only weakly inhibited by buoyancy shut-down. The high salinity signal we observed is easy to measure, so that our repeated measurements provide more detail and much more replication than had previously existed elsewhere. We thus shed new light on the scale and duration of the transport on the shelf and the intrusions offshore. If the bottom stress is comparable to a typical surface wind stress of magnitude 0.1 N/m^2 (or, because of buoyancy arrest, perhaps an order of magnitude less), this transport should be of comparable magnitude, 5 – 50 m^2/sec, to the very efficient process of wind-driven coastal upwelling. For example, if the bottom stress is about 0.05 N/m^2 (a rough estimate based on our observed mean flow over the outer shelf) and buoyancy shut-down is not a factor, we would expect this mechanism to flush the volume of the wide northwestern Australian shelf in about 140 days, a time long compared to either our event's duration (9 days) or the shutdown time (30–50 days).

[13] If the water is ejected offshore in a bottom boundary layer that extends alongshore for hundreds or even thousands of kilometers, we need to understand how the compensating onshore mass flux might occur. The most obvious solution to the problem is two dimensional (in the offshore-vertical plane), in which case there would be an onshore flow in the water column above the bottom boundary layer. If this flow existed, it would be too weak, of order 0.01 m/sec, for us to observe directly in the face of oceanic noise (tides, for example) in our limited current data set. This flow would also require, through geostrophy, that sea level slope downward alongshore from the northeast to the southwest. This is the opposite sense to the alongshore pressure gradient that we would expect in the absence of wind forcing with our observed equatorward flow. The other major possibility is that the offshore flux is balanced by an alongshore transport that decreases alongshore in compensation. We observe a northeastward alongshore transport off Port Hedland of about 4.5×10^5 m^3/sec during June 29–July 7. This transport could balance the offshore flux for about 450 km alongshore, or about the distance from Port Hedland to Broome.

[14] Although it is not presently clear how important our observed mechanism is globally, transport near the bottom of the water column will likely provide an efficient means for moving suspended materials, such as fine sediments and biogenic particulates, offshore so that they can ultimately settle into the deep ocean for recycling or deposition.

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